Physical impacts of climate change on landslide occurrence and related 8 adaptation

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Title: Physical impacts of climate change on landslide occurrence and related adaptation

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Abstract

This chapter provides a review on current understanding of different effects of climate change on landslides and debris flows in cold, temperate, and tropical mountains. We start with observed impacts of climate change on shallow landslides and debris flows, followed by discussions of rock-slope failures, and the physical processes that make climate an important cause and trigger of landslides. While an increase in extreme precipitation has been observed in many regions worldwide over the past decades, changes in frequency and magnitude of landslides are more difficult to identify. In high mountain regions with snow, glacier and permafrost slope stability is not only sensitive to changes in precipitation but also to changes in temperature. In the European Alps an increase of high alpine rock slope failures has been detected over the past decades and correlates to an increase in mean air temperature.

Future projections generally indicate a further increase of extreme precipitation events that are likely to go along with an increase of landslide occurrence. Seasonal variations of precipitation and earlier melt of snow imply changes of landslide seasonality. Changes in sediment supply can furthermore strongly condition debris flow frequency and magnitude as recent studies have shown.

We conclude this chapter with a case study, based on a model of a landslide early warning system, that outlines the potential and limitations of adaptation to future changes in precipitation.

17.1 Introduction

Economic losses from landslides have been rising over the past decades (Guzzetti, 2000; Petley et al., 2005; ISDR, 2009), mainly because of growing development and investment in landslide-prone areas (Schuster and Highland, 2001; Petley et al., 2007). Although changes in land use, including deforestation, have affected slope stability in developed and populated areas, the causes of the increasing landslide impacts remain poorly quantified (Petley et al., 2007). The anthropogenic overprint makes it challenging to detect the potential impacts of climate change on landslide magnitude and frequency, which determine landslide hazard (Lateltin et al., 2005). The magnitude/frequency concept and its application in risk appraisals are a useful framework for quantifying the potential impacts of climate change on slope stability. Yet, detecting systematic changes in landslide occurrence as well as accurately forecasting future impacts of climate change remains difficult. Few sufficiently detailed landslide inventories allow quantifying reliably changes in the magnitude or frequency of landslides. The period of historically documented landslide events are often too short and incomplete to provide statistically significant trends. Also, there is a strong documentation bias, with many more events documented for recent years than for earlier periods. Documentation deficits exist for all types of landslides, but particularly for high mountain regions and many developing countries. Yet, many high mountain areas offer the advantages of having a lesser human impact and in being more sensitive to climate change. There, landslide activity is controlled by a variety of factors, including geology, hydrology, topography, and climate, all of them linked through feedback mechanisms. For example, atmospheric warming drives glacier retreat and exposure of fresh glacial sediments, thus potentially affecting the frequency and magnitude of debris flows (Evans and Clague, 1994; Hewitt et al., 2008).

This chapter focuses on landslides and debris flows in cold, temperate, and tropical mountains. We start with observed impacts of climate change on shallow landslides and debris flows, followed by discussions of rock-slope failures, and the physical processes that make climate an important cause and trigger of landslides. We then consider landslide activity in the context of anticipated climate change. We review the latest studies on extreme meteorological events and the development of conceptual approaches to relating climate change to future landslide activity. We conclude with a model and a case study that assesses the effect of changes in rainfall on landslide activity and related losses and shows possible direction of adaptation within the framework of an early warning system.

17.2 Observed impacts of climate change on landslide activity

17.2.1 Effects on shallow landslides and debris flows

Precipitation is an important landslide trigger. In mountain regions with seasonal or perennial snow cover, glaciers, and permafrost, temperature is an additional factor for slope stability. Intense and prolonged rainfall saturates soils and produces high transient pore pressures (Iverson, 2000) (Fig. 17.1). Other factors include rainfall totals, intensity, and duration, as well as antecedent rainfall (Wieczorek and Glade, 2005; Sidle and Ochiai, 2006). Rapid snow melt may further reduce the stability of shallow soils (Kim et al., 2004). Numerous empirical models relate observed occurrences of shallow landslides and debris flows to rainfall intensity and duration (Caine, 1980; Larsen and Simon, 1993; Jakob and Weatherly, 2003) in order to derive critical intensity-duration thresholds (Fig. 17.2). These relationships provide first-order metrics that can be used in early warning (Keefer et al., 1987; Guzzetti et al., 2007).

Records of rainfall intensity and duration in high mountain regions are limited, reflecting the scarcity of climate stations, except for parts of the European Alps. In the Swiss Alps, for example, Zimmerman et al. (1997) reported debris flow-triggering rainfall intensities of ~30 to 60 mm/h for durations of ~1 hour. Dahal and Hasegawa (2008) found that rainfall intensity thresholds for shallow landslides in Nepal, a country with a limited meteorological network, differ substantially (Fig 17.2). Antecedent rainfall may modulate critical rainfall intensities (Kim et al., 1991; Glade, 1998), although many rainfall records have insufficient temporal resolution to test this assertion (Huggel et al., 2010a). Records of landslide occurrence and corresponding rainfall parameters often suffer from uncertainties regarding exact location and timing. Orographic effects on precipitation may not be captured adequately by a single or few rain gauges. In the Alps, for example, mid- to high-elevation areas generally receive most precipitation, whereas in the tropical Colombian Andes precipitation is at a minimum in high elevated glacierized areas with the consequence that rainfall-triggered landslides abound at lower to intermediate elevations (Huggel et al., 2010a). Reviews of intensity-duration thresholds of rainfall-triggered landslides have demonstrated an order-of-magnitude variance that depends, among others, on local climate, geology, soil characteristics, and land use. Individual precipitation thresholds further depend on the quality of the records of rainfall-triggered landslides.

While rainfall thresholds are an empirical means of assessing landslide occurrence, soil and rock mechanics provide a deterministic framework for slope stability under both dry and wet conditions. The rate of infiltration is controlled by the physical properties of the soil, notably porosity, hydraulic conductivity, pore size distribution, preferential flow networks, vegetation cover, topography, freezing, and land use (Sidle and Ochiai, 2006). Several models of subsurface flow at different scales have been developed, based in most cases on Darcy's law for unsaturated and saturated flow (Montgomery et al., 1997; Iverson, 2000; Uchida et al., 2001). In mountainous terrain, transient pore pressure variations may trigger shallow landslides and debris flows, often at or near the soil- or regolith-bedrock interface (Rickenmann and Zimmermann, 1993). Permafrost may act as a hydraulic barrier that promotes pore pressure elevation in the active layer (Fig. 17.3; Zimmermann and Haeberli, 1992; Rist and Phillips, 2005).

Piezometers and tensiometer measurements in natural and artificial soils using have demonstrated the large spatial and temporal variability in pore pressures (Iverson and LaHusen, 1989;

Harp et al., 1990; Montgomery et al., 2002; Simoni et al., 2004). Pierson (1980) found that 77% to 91% of the variation in maximum pore pressure could be explained by the amount of 24-hour rainfall, exclusive of effects of antecedent rainfall. Increases in pore pressure may follow precipitation peaks by up to several hours, depending on local soil characteristics (Montgomery et al., 2002). In contrast, Harp et al. (1990) observed that pore pressures may also decrease just before slope failure. It is generally assumed that an increase in extreme precipitation causes more frequent landslides, all other system components remaining constant. Extreme precipitation may be defined by: (1) relative thresholds such as percentiles of a statistical reference distribution; (2) absolute thresholds (in millimeters); (3) return periods of precipitation events; or (4) the severity of damage (Beniston et al., 2007; Trenberth et al., 2007). The 95th or 99th percentiles are commonly used to quantify precipitation extremes (or heavy precipitation) (Trenberth et al., 2007). Precipitation extremes increased in many parts of the world over the past several decades, although the increases differ seasonally and spatially in different regions (IPCC, 2007), (Trenberth et al., 2007), and large uncertainty remains in mountain belts. For example, Kyselỳ (2009) found an increase in heavy winter precipitation in the Czech Republic from 1961 to 2005, while Pavan et al. (2008) argued the opposite for summer precipitation in mountains of the Emilia-Romagna region of Italy between 1951 and 2004. Marengo et al. (2010) concluded that precipitation extremes increased in parts of Brazil, Argentina, northwest Peru, and Ecuador during the second half of the 20th century. Increases in rainfall have also been reported in many parts of North America and some areas of Asia, for example India and western and northern China (Krishnamurthy et al., 2009; Pryor et al., 2009). On the other hand, no significant change has been observed in other regions of these continents (Choi et al., 2009).

Rainstorms during the summers of 1987 and 2005 in the Swiss Alps and adjacent areas resulted in widespread debris-flow damage (Bezzola and Hegg, 2007). The largest debris flows – Varuna in 1987 and Guttannen in 2005 – had originated from periglacial areas and had volumes up to 0.5 million m^3 (Rickenmann and Zimmermann, 1993; Scheuner et al., 2009). The Guttannen debris flow occurred during a storm that produced about 170 mm of rainfall in 48 hours, the largest single storm precipitation since 1876 with a 100-year return period (Scheuner et al., 2009). The effect of the storm was exacerbated by a rise in the 0°C-isotherm to well above 3000 m asl, resulting in rainfall and runoff at very high elevations. In 1987, intense rainfall with totals up to 260 mm together with antecedent precipitation was important in triggering numerous debris flows (Zimmermann, 1990).

How trends in landslide occurrence are linked to these precipitation changes, awaits further study and integration of different time series. A growing number of landslide frequency data has become available, though mostly at the catchment scale. For example, debris-flow activity on a debris fan in the Valais, Swiss Alps, was higher during the 19th century than today (Stoffel et al., 2005). In a regional study at high elevations in the French Alps, Jomelli et al. (2004) found no significant change in debris-flow frequency since the 1950s (Jomelli et al., 2004). On the other hand, Pelfini and Santilli (2008) detected a slight increase in debris-flow frequency in a largely natural high mountain environment in the Italian Alps in the second half of the 20th century relative to the period 1875-2003; the peak in frequency occurred between mid-1970s and mid-1980s.

17.2.2 Effects on rock-slope failures and debris flows in high mountains

Rock-slope response to climatic fluctuations differs somewhat from that of soil or regolith slopes. The degree of rock fracturing has a strong effect on infiltration capacity, affecting the hydrostatic pressure resulting from the vertical height of the interconnected saturated zone. Piezometric measurements in rock slopes show that groundwater conditions are not necessarily hydrostatic. The spatial distribution of pressures varies strongly with rock structure (Watson et al., 2004; Willenberg et al., 2008). Displacement measurements in a rock slope in British Columbia, Canada, at an elevation of 500-800 m asl and with an annual temperature range of -25°C to 35°C, indicate a possible relation between transient pore-pressure variations and infiltration (Watson et al., 2004), on top of the annual temperature cycle. Cooling introduces deviatoric stresses and a reduction of effective normal stress that results in cracking of rock bridges, continuous loss of cohesion in the rock mass, and eventually, slip (Krähenbühl, 2004; Watson et al., 2004).

Few measurements of displacement have been made on steep frozen rock slopes in high mountains. Observations at Jungfraujoch (3580 m asl) in the Swiss Alps, show extension during periods of cooling and contraction during warming periods (Wegmann and Gudmundsson, 1999). Precipitation at such high-elevation sites is predominantly in the solid form, thus snow melt may dominate infiltration, depending on local topography, aspect, and rock fracturing. Liquid water generated by permafrost thaw modulates hydrostatic pressure, loss of bonding, and reduction of shear strength (Gruber and Haeberli, 2007). Recent field measurements at Matterhorn, Swiss Alps, have contributed to an improved understanding of the role of water infiltration from snow/ice melt , water circulation in fracture systems, permafrost thaw, and related deformation in alpine rock slopes (Hasler, 2011).

Only in few cases such as the 1988 Tschierva rock avalanche (~3100 m asl), eastern Swiss Alps, was it possible to reconstruct antecedent rainfall thanks to the proximity of a rain gauge. The record indicates exceptionally high precipitation (>100 mm) on two consecutive days two weeks before failure (Fischer et al., 2010). Temperatures were around freezing during this time, and therefore some of the precipitation was in solid form. Subsequent melting and infiltration may have compromised slope stability on top of controls such as faults, fracture systems, lithology, and permafrost. In high-mountain regions with permanent snow, glaciers, and permafrost, both long-term gradual increase in temperature and brief temperature extremes may destabilize slopes (Haeberli and Beniston, 1998; Gruber and Haeberli, 2007; Huggel et al., 2008). Transient temperature increases and associated production of melt water may trigger landslides without precipitation. Analysis of a number of large landslides and ice avalanches in Alaska, the European Alps, and the Southern Alps of New Zealand has shown a pattern of unusually high temperatures days and weeks before many failures (Huggel et al., 2010b). Although the specific processes are not understood in detail in every case, it is reasonable to assume that melt water due to high temperatures infiltrated rock slopes along fractures and joints, thawing frozen clefts by heat advection, possibly increasing hydrostatic pressures and thus reducing shear strength. Melt water can also penetrate to the base of steep glaciers and reduce their resistance to failure (Fig. 17.4). Also, a gradual increase in mean temperature can reduce shear strength through permafrost thaw.

Recent studies indicate that rock-slope failures increased significantly in numbers; both for large failures (>10^5 m^3) over the central European Alps (Fischer, 2010) as well as for smaller ones (10^2-10^4 m^3) on the local scale (Ravanel and Deline, 2011). This increase coincides with a marked temperature increase and with an increase of temperature extremes. The IPCC AR4 concluded that most of Earth's land regions saw an increase in high temperatures, expressed as the 90th or 95th percentiles of the long-term record, over the past 50 to 100 years (Trenberth et al., 2007). In Europe, the frequency of hot days has almost tripled over the period 1880-2005 (Della-Marta et al., 2007), and Ding et al. (2007) and Kunkel et al.(2008) found both a strong increase of heat waves since the 1960s in, respectively China, and the United States. However, studies specifically targeted at temperature trends in high-mountain regions are rare so that any causal linkage to landslide occurrence requires further study.

17.3 Projected impacts of climate change on landslide activity

Precipitation and temperature extremes may contribute to destabilizing slopes and potentially triggering landslides (Fig. 17.4), but also affect boundary conditions such as topography. Local topography can change significantly over years to decades due to shrinkage of glaciers and landslides (Holm et al., 2004; Geertsema et al, 2006; Paul and Haeberli, 2008; Hewitt et al., 2008). Temperature and precipitation, however, change on much shorter timescales. The IPCC AR4 concluded that both, extreme precipitation and temperature events, will likely increase in frequency in the 21st century, but uncertainties are large, especially for precipitation (Meehl et al., 2007), though General Circulation Models (GCM) and Regional Climate Models (RCM) help circumvent these shortcomings (Hawkins and Sutton, 2010). At the global scale, Kharin et al. (2007) forecast that return periods of extreme temperature and precipitation events will be half as long by around the middle of the 21st century as compared to late-20th century values.

Several studies have applied statistical and dynamical downscaling to project extreme precipitation in areas of complex topography. Such downscaling of GCMs for North America indicates a strong increase of heavy precipitation over the south and central United States during the second half of the 21st century, but a decrease in extreme precipitation over the Canadian prairies (Wang and Zhang, 2008). An increase in precipitation extremes is also forecast for northern and central Europe, with a possible decrease in southern Europe (Beniston et al., 2007; Schmidli et al., 2007). These trends are consistent with results of RCM results from areas in Europe (Kyselỳ and Beranová, 2009). Similarly, an increase extreme precipitation is forecast for the period 2071-2100 for most of southeast South America and western Amazonia (Marengo et al., 2009). For the occurrence of landslides in mountain areas, the season of heavy precipitation matters, for instance due to snow effects. Several models show a projected increase in extreme precipitation in winter for northern and parts of central Europe (Christensen et al., 2007; Kyselỳ and Beranová, 2009), as well as increases in extreme precipitation in winter, spring, and fall for the UK (Buonomo et al., 2007; Fowler and Ekström, 2009).

Periods with temperatures in the uppermost percentiles will be more intense, more frequent, and longer lasting in the future (Meehl et al., 2007). Daily minimum temperatures are forecast to increase faster than daily maximum temperatures, which would lead to a decrease in the diurnal temperature range, with consequent effects on weathering processes in mountains. Important progress in regional climate modeling in Europe has been made with the ENSEMBLES project that used a large number of RCMs with identical boundary conditions over Europe (van der Linden and Mitchell, 2009). The RCMs were run with 25- or 50-km horizontal resolution for the period 1951-2050 (some until 2100), assuming an greenhouse gas emission scenario of rapid economic growth (SRES A1B) (Nakicenovic and Swart, 2000). In one of the few studies focusing on temperature extremes in high mountains, Huggel et al. (2010b) used ENSEMBLE climate model data to analyse the frequency of 5-day to 30-day positive temperature anomalies with pronounced melting (above 5°C air temperature) at high-elevation sites. They found that such unusually warm events will be 1.5-4 times more frequent, and in some models up to 10 times more frequent, by 2050 as compared to 1951-2000. These events have implications for rock-slope and glacier stability, because the latter also produce large amounts of melt water.

Projecting a gradual, long-term increase in mean temperature is generally more robust than attempts to forecast future temperature extremes, although in the former case the magnitude of the temperature increase carries considerable uncertainty because of different climate models and emission scenarios. In Europe, mean temperatures are likely to increase more than the global average, with warming to be strongest in winter over Northern Europe and in summer over Southern Europe (Christensen et al., 2007). Over the Alps, the warming may be +2°C ±1°C by 2050, and +2.5-3°C ±1.5°C by 2070, compared to 1990 (OcCC, 2007). In the Andes, temperatures may rise by 3.5°C ±1.5°C by 2071-2100 with respect to 1961-1990 (Urrutia and Vuille, 2009). Similar projections have been made for North America, with a mean warming of about 3-4°C by 2080-2099 relative to 1980-1999, but with potential extreme winter warming over Alaska of up to 10°C (Christensen et al., 2007).

The potential effects of warming of this magnitude on landslide activity are difficult to assess, as they involve multiple feedback processes. Some studies have attempted to link semi-empirical and physically based slope-stability models with downscaled climate model output (Buma, 2000; Collison et al., 2000; Schmidt and Glade, 2003; Bathurst et al., 2005), but uncertainties about future rainfall intensities (Parry et al., 2007) and site-specific conditions reduce confidence in forecasts of future landslide frequency. Malet et al. (2007) assessed changes in slope stability in the French Alps towards the end of the 21st century, using downscaled climate models together with a hydrological and a slope stability model, highlighting the role of reduced snow cover. Jakob and Lambert (2009) projected changes in rainfall in the Coast Mountains of British Columbia towards the end of the 21st century using a number of climate models, applying a statistical rainfall intensity-duration curve for landslide-triggering storms to analyse changes in landslide frequency (Fig. 17.5).

Possible effects of snow-line change on slope stability in mountainous terrain are shown in Figure 17.6, for a past reference period and for an arbitratrary scenario around 2050 when glaciers will have further retreated and the snow line has risen by about 300 m vertically. Shallow landslides and debris flows commonly initiate below the seasonal snow line and in glacier forefields (Rickenmann and Zimmermann, 1993; Evans and Clague, 1994; Haeberli and Beniston, 1998) due to abundant and poorly consolidated sediment in proglacial environments, and strong infiltration and potential saturation of sediment by snow melt. Locally intense rainfall at high elevations, in combination with a high snowline, favour shallow landslides and debris flows (Rickenmann and Zimmermann, 1993; Bezzola and Hegg, 2007; Scheuner et al., 2009).

Taking into account the seasonal frequency of such events, a scenario characterised by sole changes in temperature (scenario A in Fig. 17.7) and concomitant extension of periods with higher snow line may involve (1) an extension of the "landslide season", especially in spring; (2) a shift in the peak of landslide activity to early summer; and (3) a slightly higher landslide frequency in winter, depending on snowline and site elevation. The cumulative frequency of landslides may remain approximately similar, although system feedbacks make projections difficult. For a scenario with combined increase in mean temperature and precipitation intensity (scenario B in Fig. 17.7), one may envision a similar seasonal landslide frequency distribution as in scenario A, but with higher frequency of landslide occurrence. More reliable future assessments of extreme precipitation variability depend on climate model improvement, such that seasonal trends in these curves may be specified eventually (Kyselỳ and Beranová, 2009).

In addition to temperature and rainfall intensity, sediment supply and land-use are important determinants of landslide frequency and magnitude. Observations at several sites in the Swiss Alps indicate that sediment supply can change significantly due to permafrost degradation in rock and scree slopes. In the Valais in the southern Swiss Alps, accelerating rock glaciers between the 1950s and 1990s delivered more sediment into debris flow channels by a factor of 5 or more (Lugon and Stoffel, 2010). The average velocity of rock glaciers is typically <1 ma^{-1}, but ground-based monitoring and remote-sensing since 2002 have revealed an acceleration of rock glaciers to 4 ma^{-1}, and in exceptional case to as high as 15 ma^{-1} (Delaloye et al., 2008; Roer et al., 2008). Similar increases in deformation rates, also associated with landslides, have been reported from various parts of the Alps and are well correlated to increases in mean annual air temperature (Kääb et al., 2007; Roer et al., 2008). As a consequence, debris-flow magnitude and frequency from those deposits may change because of these increased rates of sediment availability.

A possible model case for a warmer future could be the situation at Ritzlihorn – Guttannen, central Swiss Alps. Strong rock fall activity commenced from the frozen north face of Ritzlihorn in 2009, likely caused by warming and degrading permafrost. Rock-fall and avalanche debris accumulates at the apex of the large Holocene debris fan that supports the community of Guttannen (Fig. 17.8). The apex has become an initiation zone for debris flows during rainstorms, and sediment saturation from avalanche-snow melt (D. Tobler, pers. Communication, 2011). Debris flows entrain material along the fan channel, episodically obstructing the trunk Aare River. Just before reaching the river, debris flows pass over a gallery of an important highway and a transnational gas pipeline. Both lifelines were interrupted and heavily damaged by debris flows in 2009 and 2010. Unprecedented rock-fall activity from steep thawing slopes, inducing equally unprecedented debris flows at Ritzlihorn – Guttannen may be a forerunner of future conditions in similar steep glacierized watersheds.

17.4 Adaptation to effects of climate change on slope stability

17.4.1 Modeling precipitation change and effects on landslide occurrence

Historical climate records and GCMs/RCMs deliver plausible scenarios of future climate change. Any single model and projection, however, has too much uncertainty to be useful, especially for rainfall forecasts to assess landslide activity and impacts on environment and society. Forecasting changes in landslide activity must explicitly account for the spatial and temporal distribution of rainfall. In some landslide-prone regions with short, but heavy rainfall, measurements of rainfall intensity over 10-15 minutes are required to derive adequate threshold values for triggering landslides. Such detail is usually neither available nor reliable enough for long-term projections, even when boundary conditions produced by a GCM are refined in a regional climate model that is specifically calibrated for a particular region or area. To overcome this shortcoming, we can use simplified modeling techniques together with baseline historical records to estimate future trends in precipitation. Historical measurements are scaled based on projected annual, semi-annual, or seasonal precipitation change. We illustrate the application of this approach with a case study in Colombia. The study area is the Combeima valley in the tropical Cordillera Central (~4°30’ N, 75°W). Landslides from the slopes of the vally, primarily triggered by heavy precipitation, are frequent and have caused hundreds of fatalities and extensive damage to the currently ~5500 people living there (Fig. 17.1; Huggel et al., 2010a). Landslide initiation zones lie in an elevation range of ~1500-3000 m asl and thus well below snowfall: the 0°C-isotherm lies at ~4800 to 4900 m asl. Projected precipitation increases for central Colombia, including our study area, are, respectively, 10-20% and 5-10% for December-February and June-August for 2090-2095 with reference to 1980-1999 (Christensen et al., 2007). For extreme precipitation events, however, projections for the Cordillera Central are controversial. Some regional climate model studies forecast an increase in these events, whereas others predict a decrease (Marengo et al., 2009). The differences in the model outputs mean that a full range of possible scenarios, with ±20% precipitation, must be considered.

To assess the effects of precipitation change scenarios on landslide occurrence and damage, we applied a recently developed landslide early warning system (LS-EWS) model (Huggel et al., 2010a). For the baseline, we used ERA-40 reanalysis precipitation data with 6-hour resolution (Uppala et al., 2005) for the period 1991-2000 from the one grid cell closest to the Combeima valley. The model applies an empirical rainfall intensity-duration threshold to simulate landslide occurrence and subsequent evacuation in case a landslide is simulated. Monte Carlo simulations were used to model a rainfall measurement error that translates into a range of evacuation and damage scenarios within a LS-EWS scheme derived from local conditions in the Combeima valley (Huggel et al., 2010a, for a more detailed description of the model). Precipitation change scenarios were run on the ERA-40 reference data. However, a single value for the precipitation change likely produces a biased damage assessment process, because it does not provide any information on the change in magnitude and frequency of heavy rainfalls. Thus, we defined scenarios with special emphasis on heavy rainfalls, together with simple scenarios where all the rainfall intensities are multiplied by the same scaling coefficient independent of the rainfall intensity.

According to that, we have fixed a scaling parameter δ and compared two scenarios: scenario 1: all rainfalls are scaled equally by a parameter $(1+\delta)$; and scenario 2: heavy rainfalls are scaled by a parameter (1+δ) and small ones are scaled by a parameter (1-δ), so that heavy rainfalls have even higher intensity and low-intensity rainfalls are even less intense, with the total amount of yearly precipitation remaining constant. For our dataset, the triggering threshold, together with δ in the range of ±20%, the two scenarios deliver the same results in terms of the number of landslides and their expected damage (Fig. 17.9). This outcome is not surprising because landslides are triggered by extreme precipitation and both scenarios modify extreme rainfall in the initial dataset identically: a relatively small positive value of δ does not transform small rainfalls into big ones in either scenario. The model is dimensionless and spatially not explicit such that atmospheric moisture dynamics do not affect the model performance.

17.4.2 Adaptation of a landslide early-warning system to changing precipitation

We applied the LS-EWS model to assess damage under different precipitation scenarios. The purpose of a LS-EWS is to issue reliable warnings to allow people in the endangered area to evacuate to safety while damages to buildings cannot be avoided. To evaluate EWS performance in a changing climate, it is necessary to discriminate between losses that can be prevented by such a system and those that cannot. For that reason, the losses estimated here (including those shown in Fig. 17.9) are restricted to evacuation costs, injuries, and fatalities, and exclude damage to infrastructure. Basically, the EWS performs better if the losses are reduced. The capacity of an LS-EWS to adjust to changes in precipitation comprises internal and external elements. Internal elements involve rules relating to warning and evacuation, for example the circumstances that lead to issuing of an evacuation order. External elements involve changes in the key functional components of the EWS, for example improvements in spatio-temporal resolution of rainfall measurements. Figure 17.10 illustrates the internal and external adaptation capacities of the LS-EWS of our case study in Colombia. The negative values of δ represent a reduction in precipitation and lead to a smaller number of landslides and consequently reduced losses (down to zero for δ = –0.2; Fig. 17.10, left panel). For a fixed rainfall measurement error (RME), only the evacuation threshold is adjusted in response to a change in precipitation. The sensitivity of the LS-EWS relevant losses to the evacuation threshold is shown in Figure 17.10 (right panel). The relative loss reduction of the adjusted EWS versus the unadjusted EWS is expressed as: $(L_u - L_a)/L_u$, where L_u are expected losses for unadjusted EWS and L_a are those for adjusted EWS. The relative loss reduction is moderate to fairly high (10-25%) for a combination of moderate precipitation change (δ=0.05, δ=0.1) and high-precision EWS (RME<0.15). For a less precise EWS (RME>0.2), the efficiency gain from the adjusted evacuation threshold is negligible and comparable to the graph fluctuations of ~5% that are inherent to Monte-Carlo simulation procedures of the model. The efficiency gain is also ~5% for a larger increase in precipitation (δ=0.2) uniformly for all RMEs, indicating nearly no sensitivity of the EWS to the evacuation threshold adjustment for high precipitation changes.

Figure 17.10 (left panel) shows the potential and limitations of RME improvement for EWS adaptation purposes. From a baseline scenario under present conditions (δ=0) with a high rainfall measurement error (RME=0.3) losses can be slightly reduced for a climate change scenario with strong precipitation changes (δ=0.2) but with substantially improved rainfall monitoring capacities (RME=0.05). Our modeling results indicate that the RME improvement representing external adaptation capacity seems to have a greater potential for successfully improving an EWS than an adjustment in the evacuation threshold alone. Figure 17.10 (left panel) furthermore shows that the relative loss reduction of an RME-improved EWS(RME=0.05) over the initial EWS (RME=0.3), even for an extreme precipitation increase (δ=0.2), may be about as large as (2.7-1.1)/2.7 ≈ 60%, which is substantially larger than the loss reduction achieved with an adjustment in the threshold alone (Figure 17.10, right panel).

Several considerations of LS-EWS in general, and of this approach of assessing EWS adaptation capacity in particular, should be taken into account. Increased precipitation most likely will result in an increased number of landslides under otherwise unchanged conditions. Structural mitigation measures such as dams or passive avoidance such as relocation of endangered assets, including population, may be alternative or complementary options to EWS. Furthermore, the assumption that the system is stationary and, for example, the triggering threshold does not change is oversimplified. A major difficulty with LS-EWS is that rainfall events with intensity-duration relations exceeding the threshold may not trigger landslides but require issuing an alarm. Recent studies using different rainfall exceedance probabilities are promising, but do not fully address the problem of non-landslide rainfall events (Brunetti et al., 2010). Soil mechanics models may contribute adding even more uncertainties for LS-EWS, given the high spatial variability of soil and rainfall conditions. Finally, human response to warnings remains a critical issue, and the willingness of people to evacuate may decrease after false alarms (Dash and Gladwin, 2007). Communication of the early warning procedures and uncertainties is crucial (Dow and Cutter, 1998), though clearly more research is needed to reasonable implement such aspects in a numerical model.

17.5 Conclusions

Different types of landslides respond differently to changes in precipitation and temperature. Despite growing number of data it is unclear whether an increase in extreme rainfall events is likely to increase the occurrence of shallow landslides and debris flows, assuming otherwise constant conditions. Even though extreme precipitation events have increased in many regions of the world over the past several decades, significant, let alone systematic, changes in landslide magnitude-frequency distributions remain to be detected. Reasons for this include incomplete and biased documentation of events that limit extraction of climatic signals, and overprints of land-use changes. In some high-mountain regions such as the central European Alps, observed large rock slope failures have increased during the past

100 years. Although the mechanisms are not completely understood, rapid and large glacier shrinkage and permafrost degradation are likely to have had a considerable influence.

Reliable forecasts of future climatically driven changes in landslide activity must overcome a number of challenges. Hydrological and slope stability modeling demonstrates the strong influence of local site conditions, taking advantage of scenarios in which landslides may become more or less frequent. Most climate models, however, indicate an increase in the frequency of extreme precipitation and temperature events during the 21st century. It is likely that landslide frequency and magnitude will also be influenced by such trends, especially in high mountains where changes in sediment supply can be significant and play a critical role in making more material available for slope failure.

The necessity for landslide early warning systems is rising. Changing rainfall patterns, however, pose an additional challenge to the development of robust and reliable warning systems. Studies indicate that damages related to landslides may rise considerably with higher rainfall in the future if EWS are not adjusted. Adaptation involving improved rainfall observations may be a better option than internally adjusting evacuation procedures, such as changing evacuation thresholds for rainfall. On a more general level, thus, our studies support the call for improved environmental monitoring as one necessary and effective adaptation strategy for future climate change.

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Figure 17.1: Two shallow landslides in the Combeima valley, Colombia, that occurred in 2006 (right panel) and 2009 (left panel). The image on the left shows the source zone of a shallow, unchannelized landslide that initiated ona steep (~35°), cultivated slope. The landslide on the right is much larger and was channelized, with deposit heights of 4-5 m (persons and car for scale). (Photos by C. Huggel.)

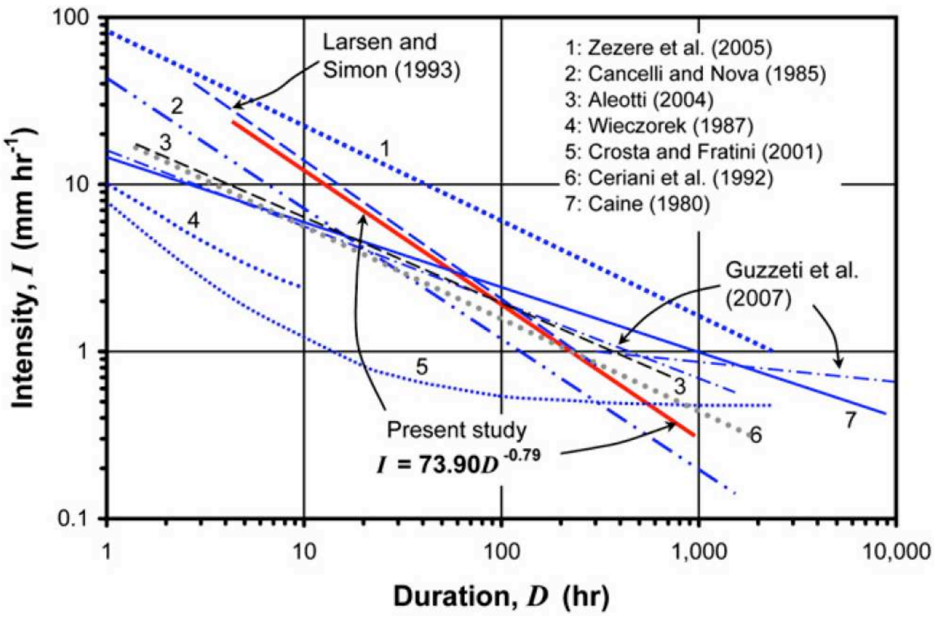

Figure 17.2: Published landslide-triggering thresholds of rainfall intensity and duration thresholds. The thresholds can differby an order-of-magnitude, depending on climate, hydrology, topography, geology, land cover, land-use, and other factors (from Dahal and Hasegawa, 2008).

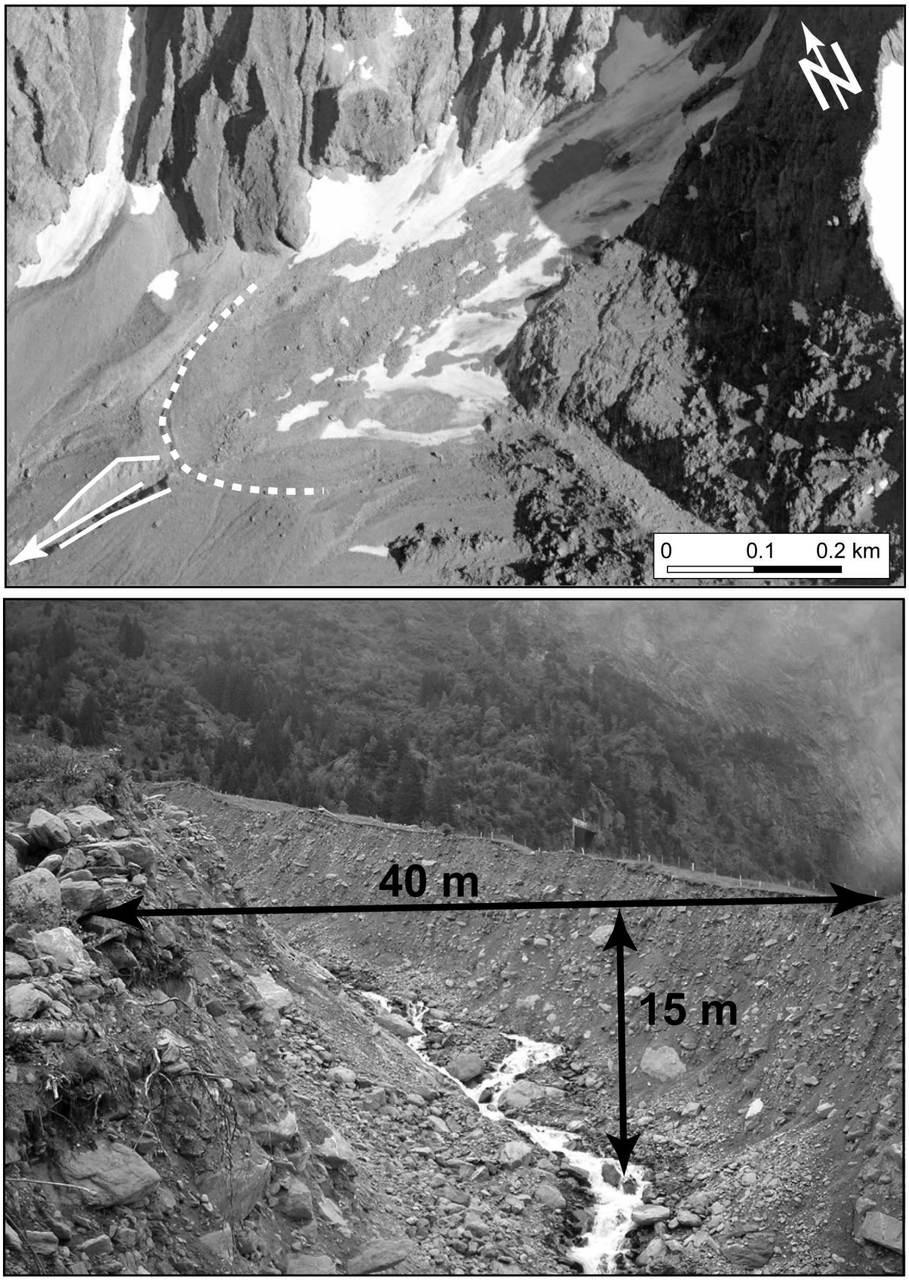

Figure 17.3: Source (upper image) and eroded channel (lower image) of the 22 August 2005 Rotlaui debris flow near Guttannen in the central Swiss Alps. The dashed shite line on the aerial image delineates a body of permanently frozen sediment, formerly covered by Homad Glacier. The arrow indicates the initiation zone of the debris flow (~2400 m asl). A significant percentage of the debris flow volume was entrained on the Holocene debris fan between ~1100 and 1300 m asl (lower image). The debris flow was the largest (~500,000 m^3) in Switzerland in at least ~20 years.

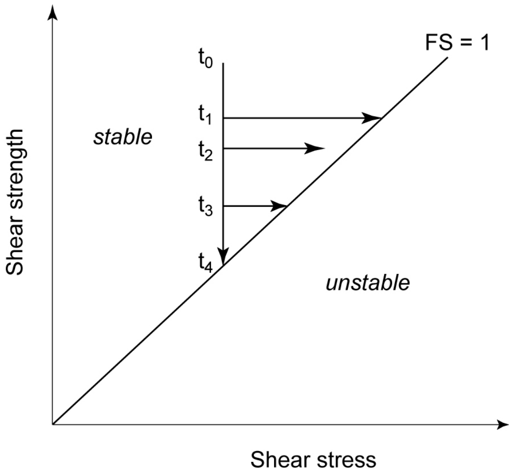

Figure 17.4: Simplified scheme illustrating a theoretical slope history with a long-term gradual decrease in slope strength due, for example, to permafrost degradation (vertical axis) and short-term increases in shear stress due, for example, to an earthquake or heavy rainfall (horizontal axis). At different times, short-term events may increase shear stress to the threshold of instability (t_1 and t_3), whereas in others the threshold is not reached (t_2). The factor of safety (FS) basically distinguishes between stable and unstable slope conditions.

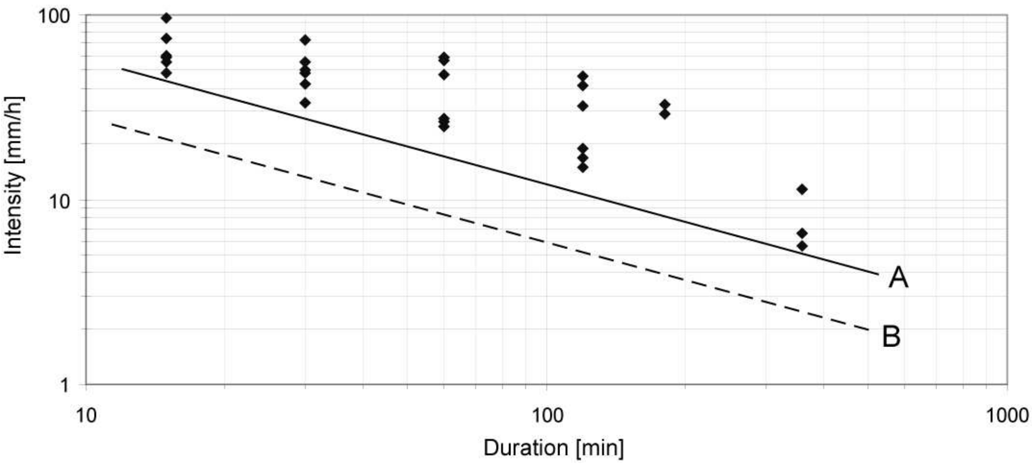

Figure 17.5: Rainfall intensity – duration plot for a number of landslides in the Cordillera Central in Colombia. Line A is a threshold function for conditions of the past few years. Line B is a scenario threshold function for possible future conditions with more frequent landslides but the same rainfall intensities. Currently it is not clear whether more intense or more frequent rainfall would shift the threshold from A to B (or any similar shift). Other processes may have a more important effect, for example a decrease in vegetation cover.

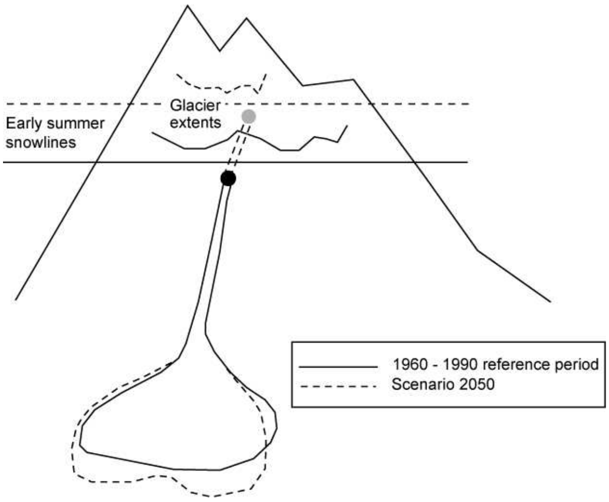

Figure 17.6:Sketch showing changes in conditions favouring early summer debris flows on a snow- and glacier-clad mountain between a reference period (in this example 1960-1990) and the future (2050). Black and grey dots indicate the points of debris flow initiation for, respectively, the reference period and the future. The limit of glacier and seasonal snow cover shifts to higher elevations, potentially raising debris flow initiation locations. The corresponding increase in potential energy of debris flows may lead to longer runouts, as shown in the sketch.

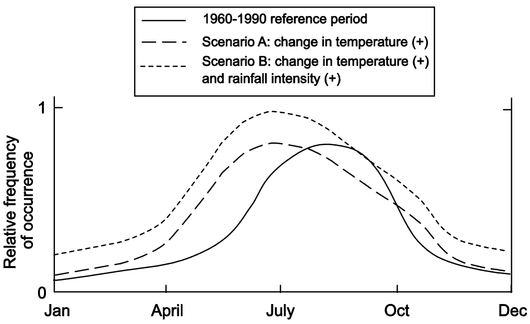

Figure 17.7: Relative frequency of occurrence of shallow landslides and debris flows in mountain areas with seasonal snow cover. For the reference period (1960-1990), landslides are most frequent in early summer when snowline recedes and meltwater runoff is high. In Scenario A, with a temperature increase, snow melt begins earlier in the year and, accordingly, shallow landslides and debris flows are more likely to occur during the spring. In addition, the landslide season is extended due to later onset of snowfall. In Scenario B, warming is accompanied by higher intensity rainfall and the frequency of landslides increases.

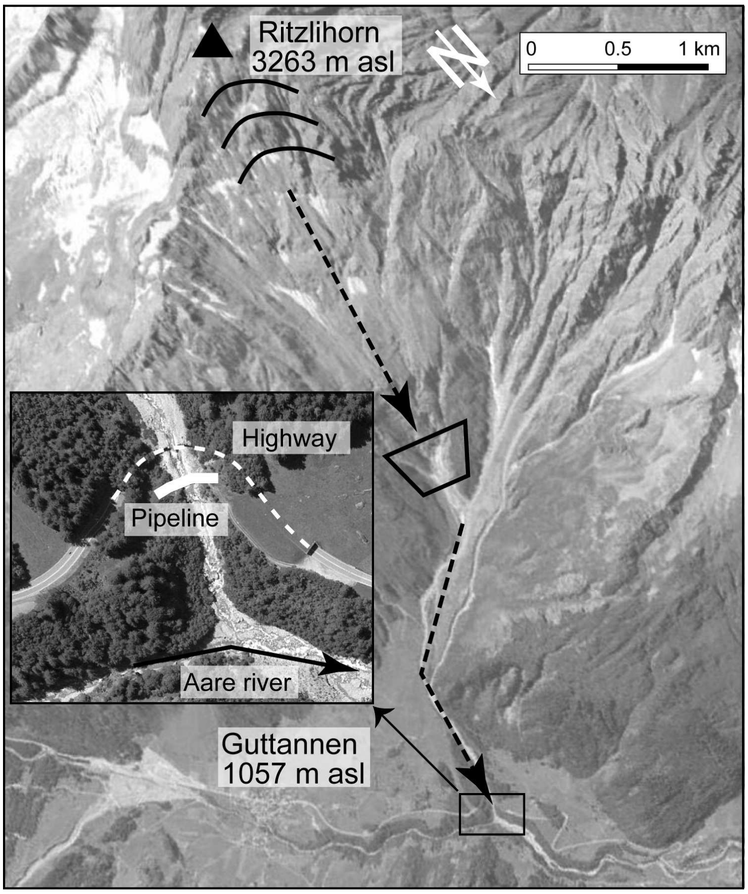

Figure 17.8: The situation at Ritzlihorn – Guttannen in the central Swiss Alps. Half circles indicate source zones of rock fall from permafrost areas on the north face of Ritzlihorn. Rock fall (upper dashed arrow) accumulates at the apex of the Holocene fan and is remobilized by debris flows (lower dashed arrow) that flow across the fan to the Aare River. The debris flows cross a highway gallery and a transnational gas pipeline(inset). (Imagery retrieved from GoogleEarth).

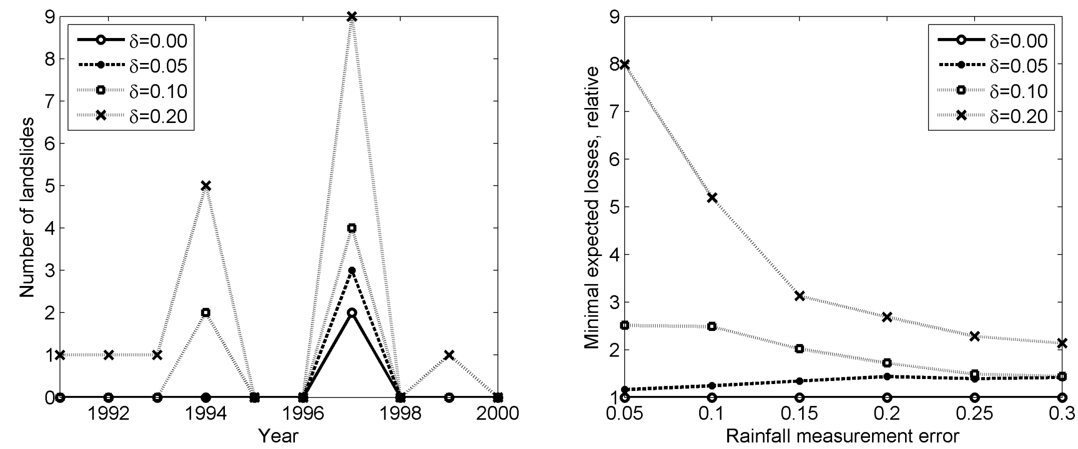

Figure 17.9: Left: Modeled number of landslides for four precipitation scenarios. All rainfall intensities were scaled equally by (1+**δ**), independently of the initial rainfall magnitude. Right: Expected damage (evacuation costs and fatalities) caused by landslides for different changes in precipitation, **δ,** relative to a baseline.

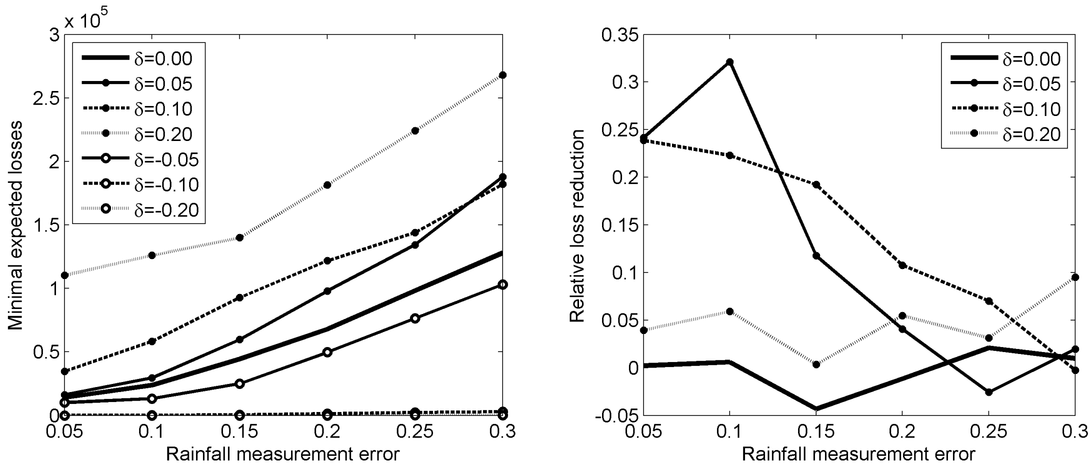

Figure 17.10: Indication of external (left) and internal (right) adaptation capacities of a landslide early warning system. Here external adaptation capacity refers to the reduction of the rainfall measurement error (RME). The internal adaptation capacity refers to the evacuation threshold adjustment for a fixed RME.